AD-A217 195

DTIC FILE COPY

(1)

REPORT DOCUMENTATION PAGE

Form Approved
OMB No. 0704-0188

1. AGENCY USE ONLY (Leave blank)	2. REPORT DATE	3. REPORT TYPE AND DATES COVERED
	31 January 1983	Final (1 December 81-30 Nov 82)

4. TITLE AND SUBTITLE

MAGNITUDE ANOMALIES AND PROPAGATION OF LOCAL PHASES

5. FUNDING NUMBERS

PE: 61101E
3291-32

6. AUTHOR(S)

B. Massinon, P. Mechler
M. Bouchon, Y. Cansi, N. Ravalau

7. PERFORMING ORGANIZATION NAME(S) AND ADDRESS(ES)

RADIOMANA- Societe Civile
27 rue Claude Bernard
75005 PARIS - FRANCE

8. PERFORMING ORGANIZATION REPORT NUMBER

AFOSR-TR- 89-1748

9. SPONSORING/MONITORING AGENCY NAME(S) AND ADDRESS(ES)

AFOSR
BLDG 410
BAFB DC 20332-6448

10. SPONSORING/MONITORING AGENCY REPORT NUMBER

AFOSR-80-0082

11. SUPPLEMENTARY NOTES

12a. DISTRIBUTION/AVAILABILITY STATEMENT

Approved for public release; distribution unlimited.

12b. DISTRIBUTION CODE

13. ABSTRACT (Maximum 200 words)

There are three main parts in this report:

- a study of magnitude anomalies in French Polynesia. A first approach gives an anomaly per station which is roughly a function of azimuth valid for all French Polynesia plus a station's constant. A more detailed study shows the influence of local structure.
- a theoretical study of generation and propagation of crustal seismic waves, Radial and vertical displacement are computed up to 500 km for various sources (quakes and explosions).
- Propagation of T waves and conversion into seismic waves at a continental slope level. We explained the rather long duration of T phases in continental station (some minutes, in French Polynesia only some 10 seconds) by conversion of water waves to seismic waves along a large area of the continemtal slope. This was verified experimentally.

14. SUBJECT TERMS	15. NUMBER OF PAGES	16. PRICE CODE
	29	

17. SECURITY CLASSIFICATION OF REPORT	18. SECURITY CLASSIFICATION OF THIS PAGE	19. SECURITY CLASSIFICATION OF ABSTRACT	20. LIMITATION OF ABSTRACT
unclassified	unclassified		

NSN 7540-01-280-5500

Standard Form 298 (890104 Draft)
Prescribed by ANSI Std. 239-18
298-01

90 01 04 091

AIR
NOT
Grant Number : 80 - 0082

(AFSC)
reviewed and
IAW AFR 190-12.
Division

MAGNITUDE ANOMALIES AND PROPAGATION OF LOCAL PHASES

Principal Investigators : B. Massinon and P. Mechler

Redactors : M. Bouchon ° Y. Cansi °° B. Massinon °°
P. Mechler °°° N. Ravaîau °°°

° Laboratoire de Géophysique Interne
Université Scientifique et Médicale de Grenoble
BP 53 X, 38041 Grenoble, France

°° Laboratoire de Détection et de Géophysique
Commissariat à l'Energie Atomique
BP 136, 92124 Montrouge Cedex, France

°°° Laboratoire de Géophysique Appliquée
Université Paris VI
T. 15 - E. 2
4 Place Jussieu, 75230 Paris Cedex 5, France

January 1983

Annual Scientific Report, 1 December 1981 - 30 November 1982

Prepared for :

ADVANCED RESEARCH PROJECTS AGENCY (DOD)

and

EUROPEAN OFFICE AEROSPACE RESEARCH AND DEVELOPMENT
LONDON - ENGLAND

Magnitude anomalies in French Polynesia

At the last DARPA meeting we presented preliminary results of our study of magnitude anomalies in French Polynesia.

The set of data consist of all quakes between January 1978 and July 1979 and recorded in at least one of the nine stations of our network in French Polynesia. This oneand a half year period was not long enough to provide us with a large enough data set to smooth our results for a large variety of azimuth. We increased our data basis by including selected events from 1975 to 1981 which occur in less seismic areas. Our final data set is made of 733 quakes among which 299 are recorded in the nine stations.

We will first, present results from our new data file which prove quite similar to those presented last year.

I - Presentation of our data set:

The detectability of the network is presented in figure 1. The cumulated number of quakes is plotted versus magnitude for both subarray of Tahiti and Rangiroa. The b factor of the asymptotic law

$$\log N = a - bm$$

valid for large magnitude is quite similar in both places: 1.68 in Tahiti and 1.65 in Rangiroa. The 50 % detectability threshold is also quite similar. Respectively 5.3 and 5.2 .

The following figures represent more precisely our data set.

Figure 2 : Geographical repartition of epicenters. This is only a seismicity map of Pacific Ocean (Tahiti beeing roughly at the center of the Pacific).

Figure 3: Number of quakes versus distance.

Figure 4: Number of quakes versus focal depth.

Figure 5: Number of quakes versus magnitude.

II - First results on magnitude anomalies.

We computed the observed magnitudes of all events in all our stations, and compared the results to the magnitudes published by USGS.

Figures 6 and 7: magnitude anomalies versus magnitude USGS for Tahiti and Rangiroa (magnitude for Tahiti means average magnitude computed from our 5 stations in Tahiti, for Rangiroa average from 4 stations in Rangiroa).

Figure 8: magnitude anomalies versus distance for Tahiti and Rangiroa.

Figure 9: magnitude anomalies versus focal depth for Tahiti and Rangiroa.

In all figures, the density of grey is proportionnal to the number of data. There is no statistically significant variation of magnitude anomalies versus one of this above parameters.

A contrario, we observed a significant dependance between azimuth of incoming wave and magnitude anomalies. This is clearly seen on:

Figure 10: Magnitude anomalies versus azimuth for Tahiti and Rangiroa and on :

Figure 11: Magnitude anomalies versus azimuth for the 5 stations of Tahiti.

Figure 12: Magnitude anomalies versus azimuth for the 4 stations of Rangiroa.

In all the nine stations of French Polynesia, we may notice that generally speaking all events east of our stations are well recorded (positive magnitude anomaly) and all events west of our stations poorly recorded.

In oder to decrease errors (both our reading errors and errors in the magnitude values published by USGS) we first decided to take average of all events from the same seismic area.

We were puzzled by the results : magnitude anomalies versus azimuth in one particular station are the sum of a constante (depending from the station) and a variation with azimuth which is the same for all polynesian stations.

It implies that there are no effects from the local structure of this island or atoll. This joint was important enough to demand a more detailed analysis.

III - Local dependance of magnitude anomalies.

A smoothing of our data on all quakes originating in the same seismic area, shows no influence of the local geology. We used E. Flinn ' s seismic zoning for this averaging process. We use only the first digit of his regionalization and so do average over a large enough number of quakes. But in that case, the size of the seismic area is too large, and so do the azimuth intervals over which we average. All possible local effects of the structure of Polynesian islands are smoothed out.

In order to study these local effects, we used in our data set, only the 299 quakes recorded in all the 9 stations.

Figure 13 gives a plot of the variation versus azimuth of the observed magnitude in Tahiti minus observed magnitude in Rangiroa. On the left hand side, a plot of all values, and on the right hand side a smoothing of variations. (The smoothing is obtained by a Bartlett triangular window, the total aperture of which is 10°).

We now notice an average effect of the structure of the island of Tahiti and of the atoll of Rangiroa.

Figures 14 and 15 try to describe in more details the effects of local stucture inside a subarray (figure 14 Tahiti, figure 15 Rangiroa).

In both figures we plotted the difference between observed magnitude in one particular station and the average magnitude of the subarray.

With this more accurate smoothing, the data lead us to show a definite effect of the local structure on the azimuthal variation of magnitude anomalies. One of our current work is to correlate this magnitude variation with what we know on the geology of the subarrays.

THE CUMULATED NUMBER OF QUAKES VERSUS MAGNITUDE

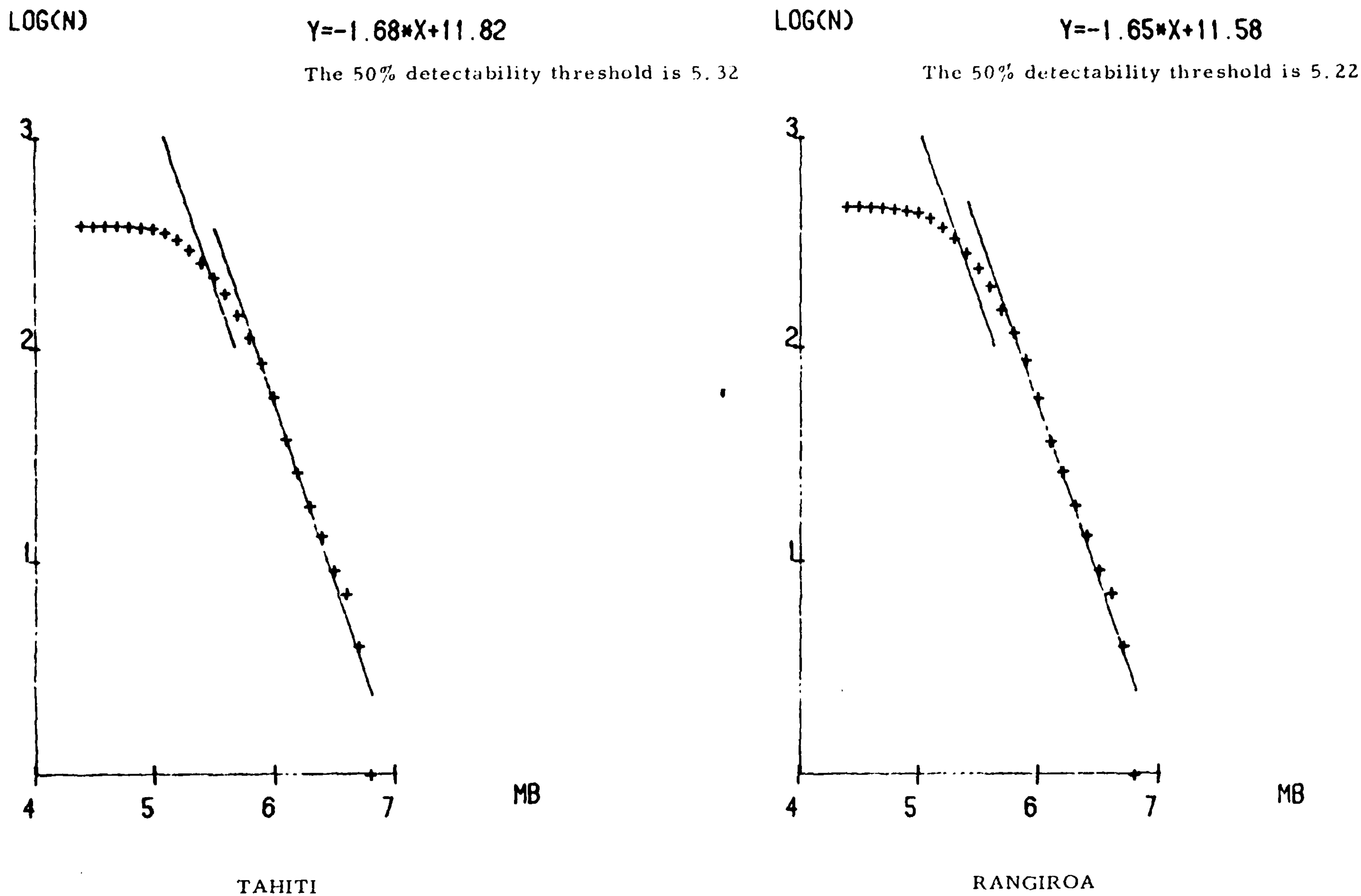


FIG.1

Geophysical repartition of epicenters

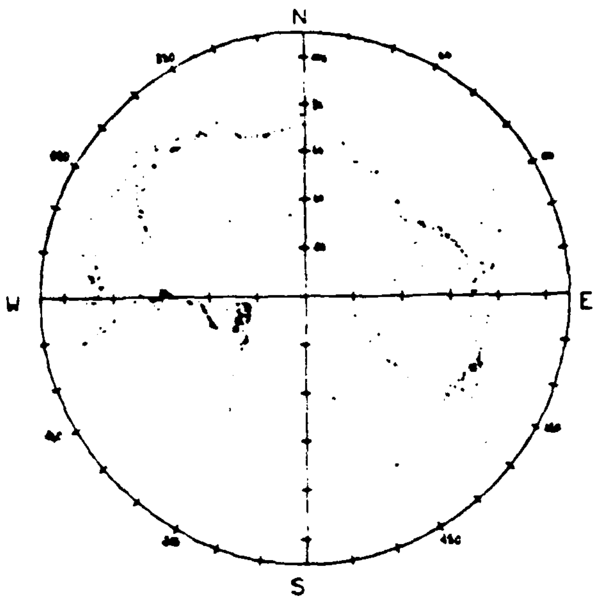


fig: 2

Number of quakes versus distance

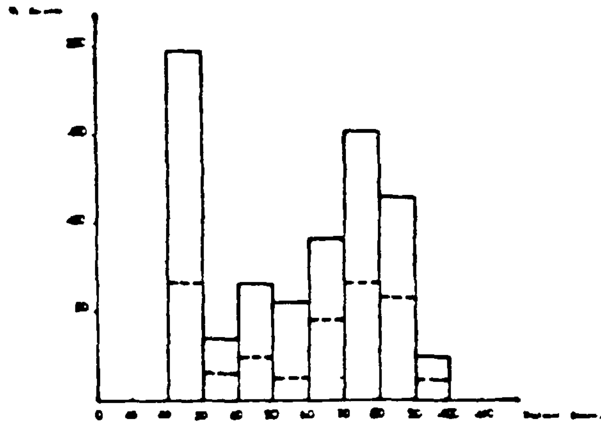


fig:3

fig: 4

Number of quakes versus focal depth

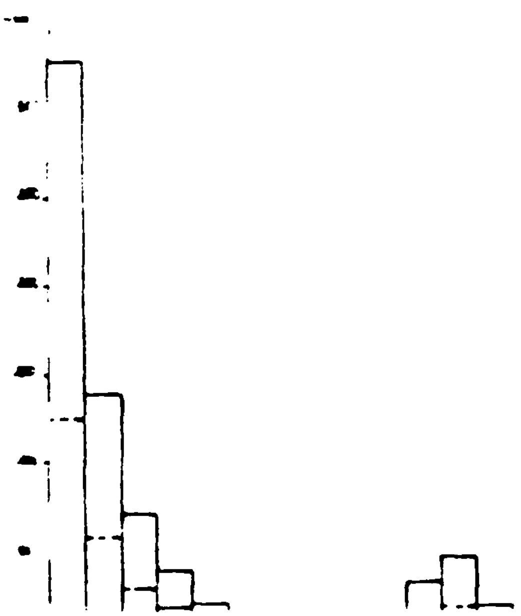
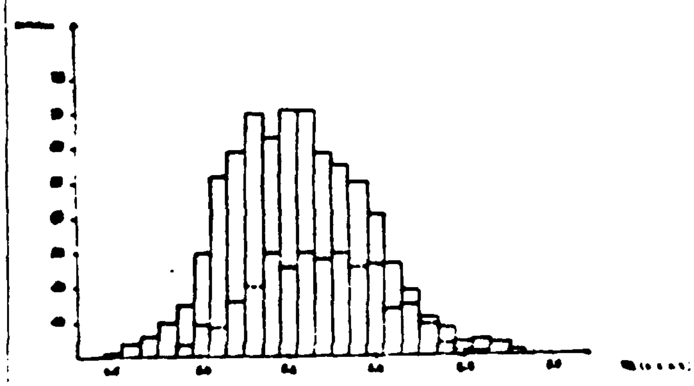


fig: 5

Number of quakes versus magnitude



Magnitude anomalies versus
magnitude USGS for Tahiti

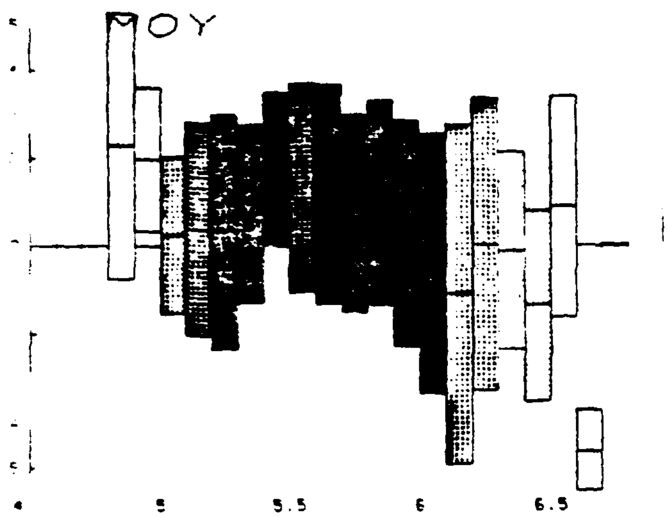


Fig:6

Magnitude anomalies versus
magnitude USGS for Rangiroa

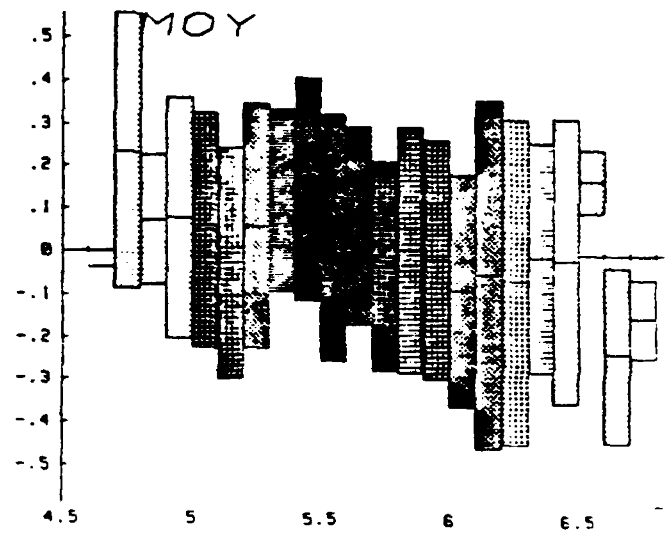
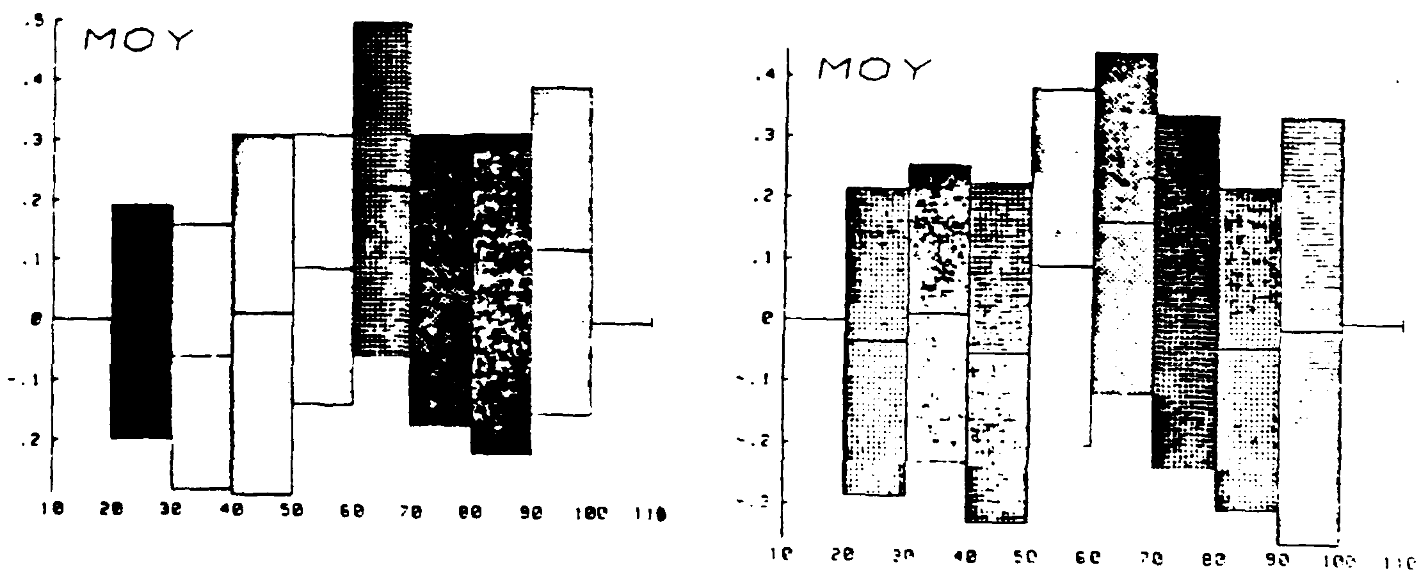


Fig:7

Magnitude anomalies versus distance.

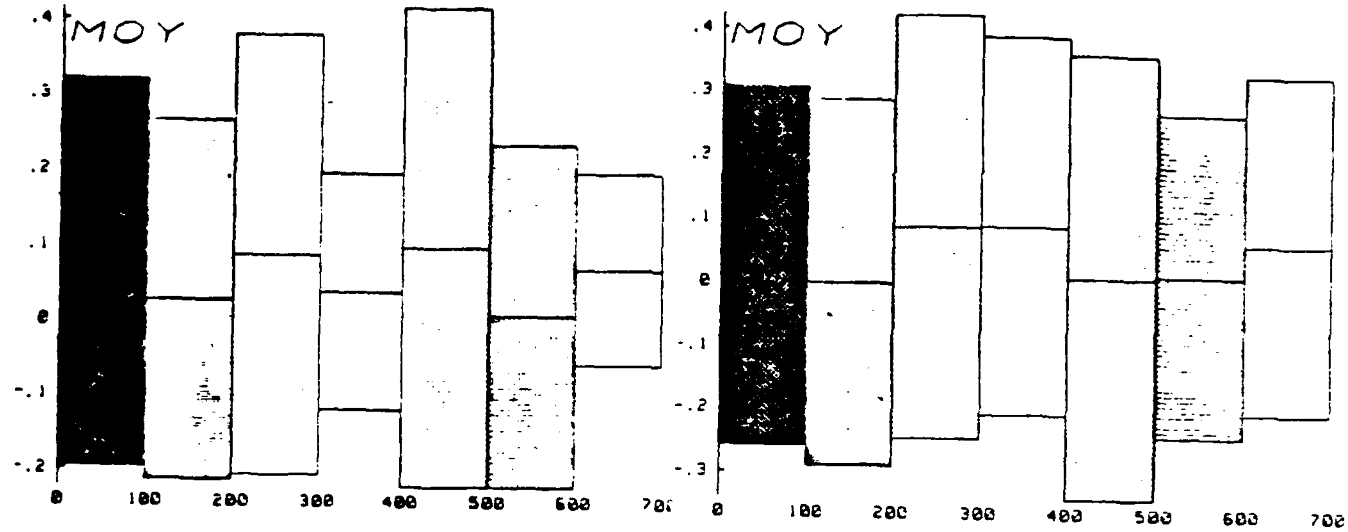


Tahiti

Rangiroa

Fig:8

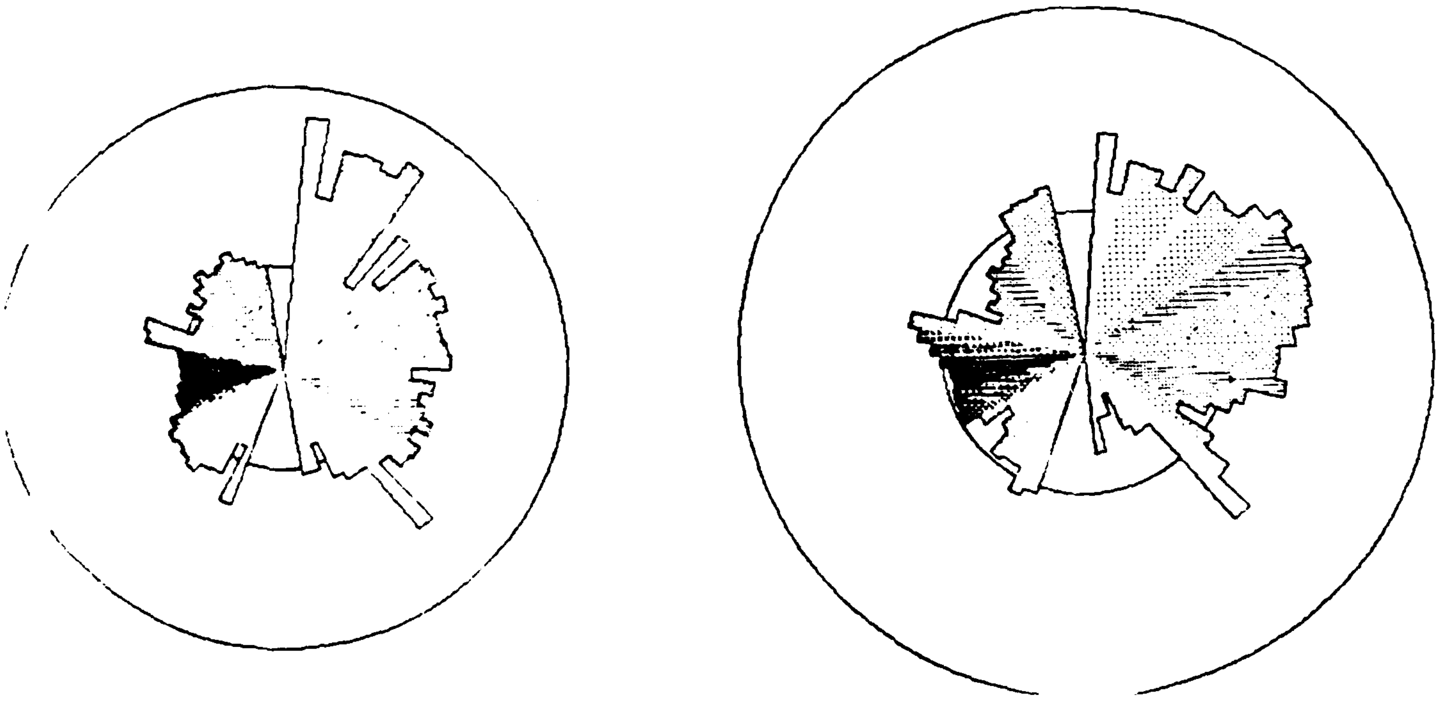
Magnitude anomalies versus focal depth



Tahiti Rangiroa

Fig:9

Magnitude anomalies versus azimuth



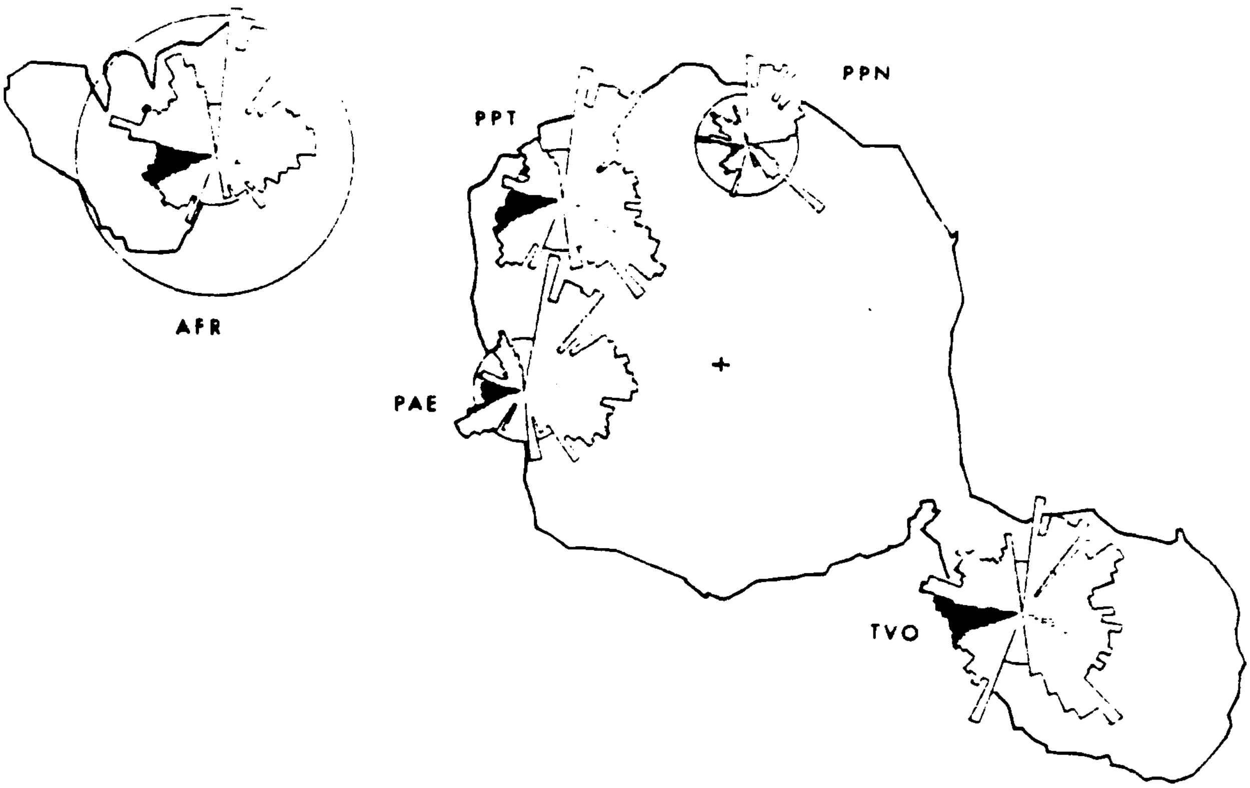
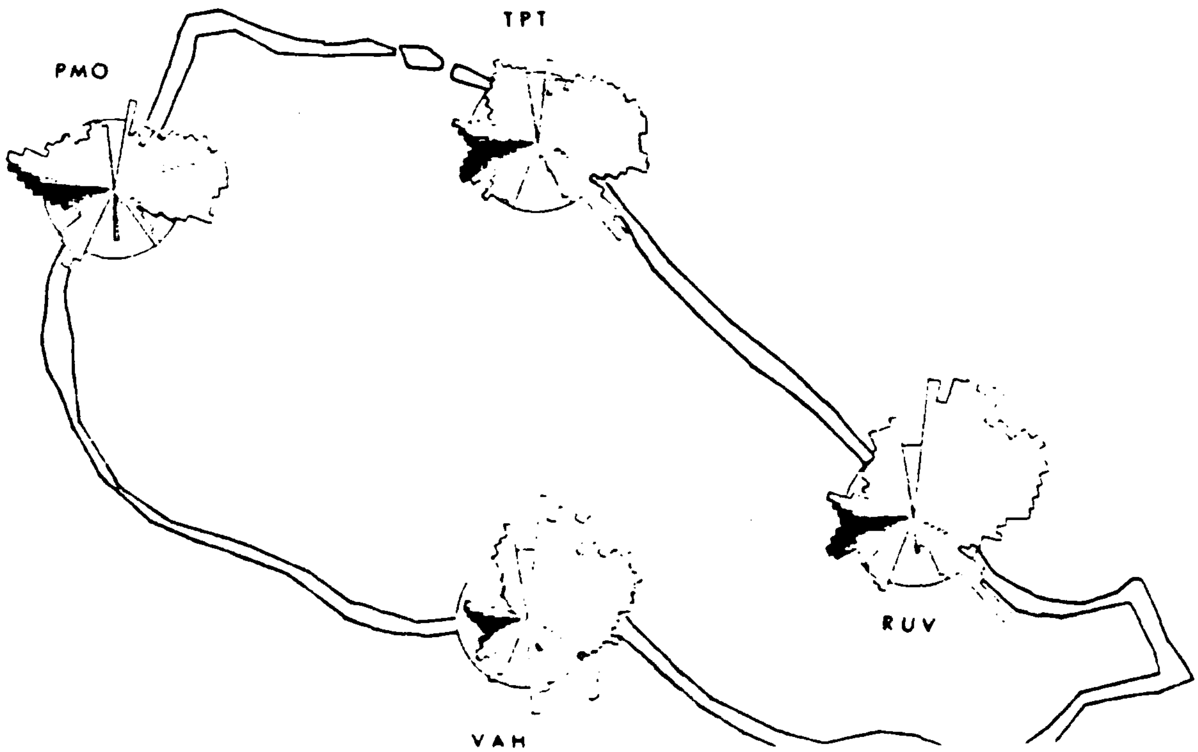


Figure 11 Tahiti : $Mb_{station} - Mb_{USGS}$ versus azimuth

Figure 12 Rangiroa : $Mb_{station} - Mb_{USGS}$ versus azimuth



MAGNITUDE IN TAHITI MINUS MAGNITUDE IN RANGIROA

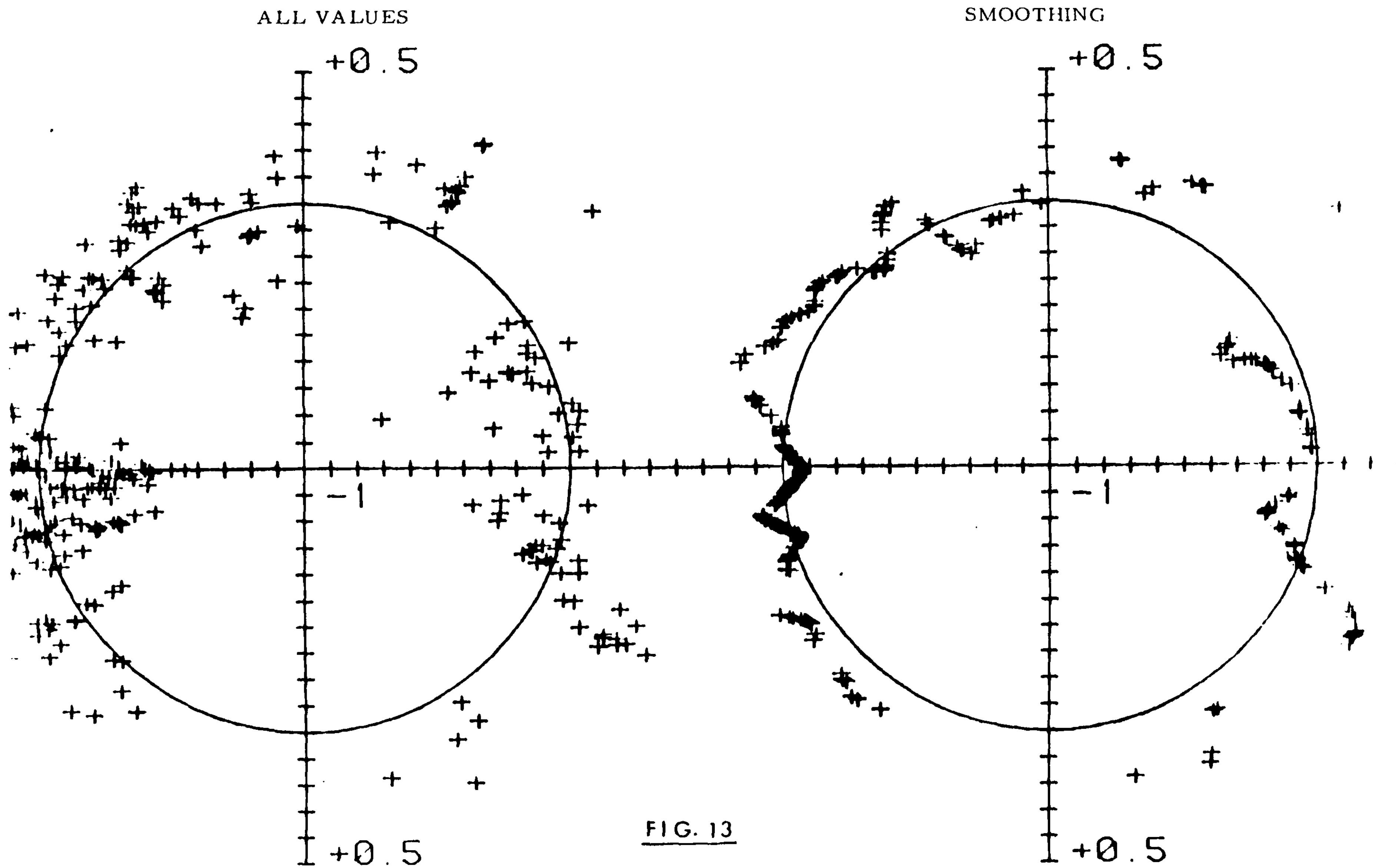


FIG. 13

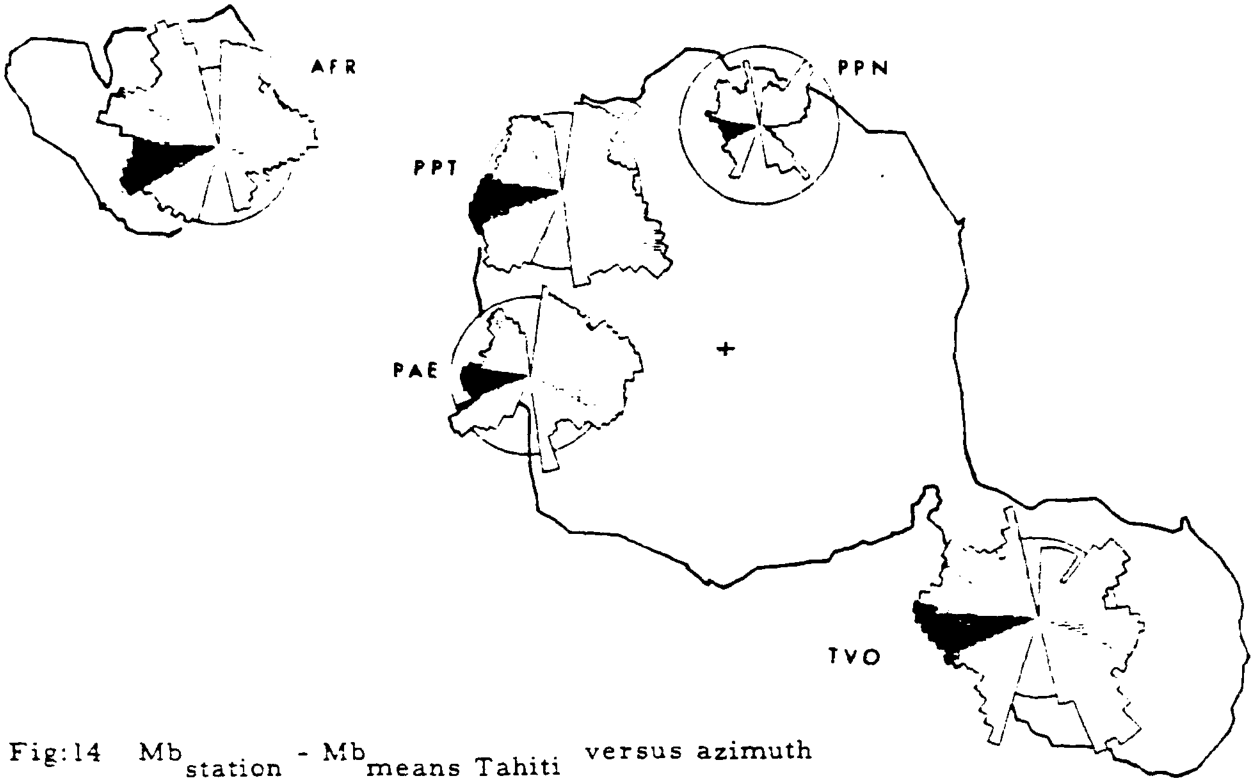
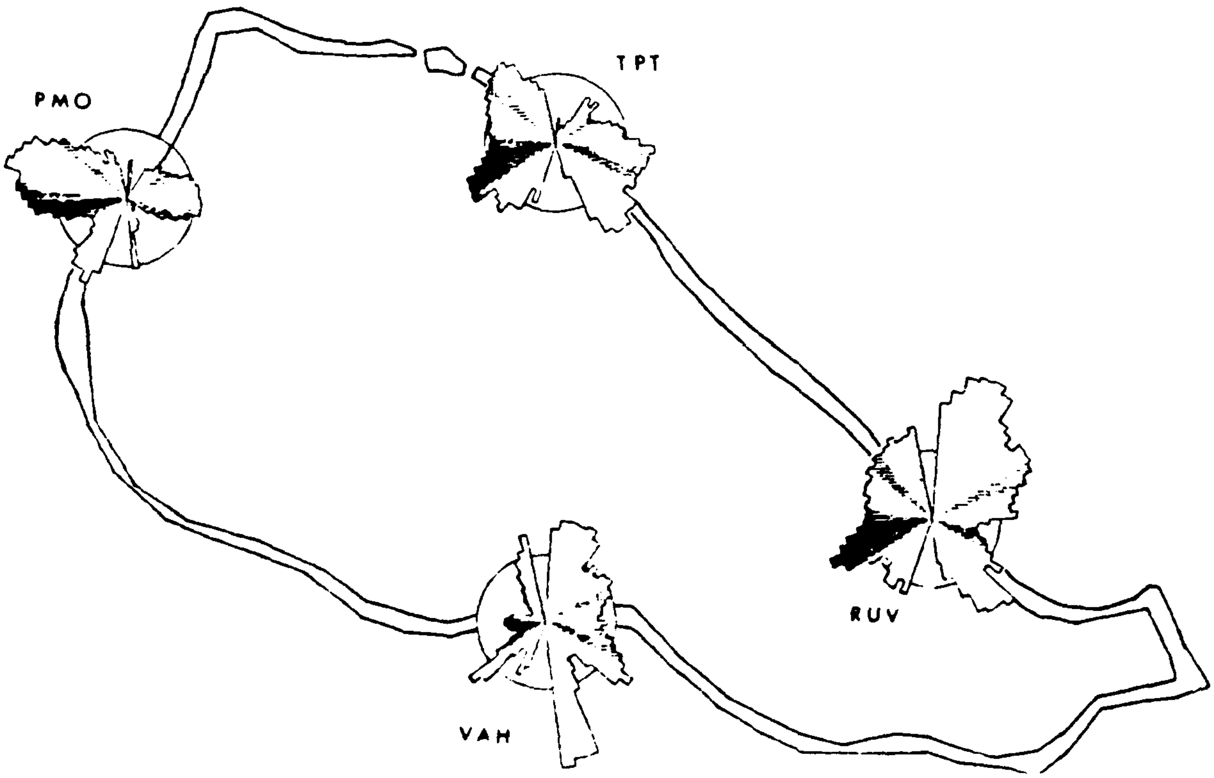


Fig:14 $Mb_{station} - Mb_{means\ Tahiti}$ versus azimuth

Fig:15 $Mb_{station} - Mb_{means\ Rangiroa}$ versus azimuth



Theorical study of generation and propagation of crustal seismic waves

It has been shown in a recent study (Bouchon 1982) that the Lg waves train is a superposition of S waves multiple - reflected through the crust and incident on the Moho with an incident angle larger than the critical reflexion angle. The energy transmitted by these waves is consequently trapped inside the crust.

Pg waves are similarly built by a superposition of reflected P waves inside the crust. But these waves beeing able to be converted in S waves through the Moho, their energy is not trapped in the crust, as efficiently as it is for Lg waves.

Synthetics for earthquakes and explosions have been generated at different epicentral distances between 50km and 500km (Fig. 1a, b, c). The source is 5km depth, the mechanism is a strike slip (horizontal), and the crustal model is a continental one currently used for France (Fig. 3). From these seismograms (3 components) it has been possible to look at the maximum amplitude decrease versus distance (Fig. 2) = attenuation for Pg phases is stronger than for Lg phases. For distances larger than 100km the three components amplitudes are decreasing similarly.

These synthetics have spectra (0-5Hz) with steepish decrease for Lg phases with $f > 2$-3Hz; Pg spectra beeing rather flat up to 5Hz (at least for the involved distances).

Source depth effect on Pg and Lg phases =

Computations of synthetics have been made for three other depths = 17km, 29km, 31km. (the Moho beeing fixed at 30km). If surface waves amplitudes are strongly affected by depth, it does not seem that Pg and Lg waves are decreasing as long as the source is kept inside the crust.

When the source is put under the Moho (z = 31km) Pg and Lg amplitudes are this time strongly decreasing, observation which confirms the crust

as a wave guide for these phases.

Attenuation due to geometrical spreading has been obtained for these distances (50-500km) = (Fig. 4)

for Pg $\alpha D^{-1,5}$

for Sg $\alpha D^{-0,85}$

source mechanism effect =

Synthetics built for two different sources =

- dip slip (fault plane with 45° dipping angle).
- strike slip.

(b th sources at 5km depth).

give very similar Lg waves trains (but the polarity signs which are opposite). Pg waves are quite different. The $\frac{Pg}{Lg}$ ratio is larger for a dip slip than for the strike slip mechanism.

Case of explosions =

We have modelled the seismic source for a nuclear explosion by a pressure applied on a spherical cavity's wall, the cavity radius beeing taken is 100m and the pressure function very similar to a pulse.

Synthetics obtained with the source placed at 5km depth, have no Lg waves (Fig. 5a, b). When decreasing the depths that is to say when taking $z = 2km$ (Fig.6a, b) and $z = 1,5km$ (Fig. 7a, b) the Lg waves train is becoming important again. These few trials pointed out the influence of depth factor on the excitation of Lg waves. For explosions, these Lg waves are certainly due to P to S waves conversion at the surface level.

That study is going to be developped by adding more earthquakes synthetics in various crust models., and comparing them to real cases. These samples will be taken in France and Africa essentially.

Figure 1a :

Radial displacement

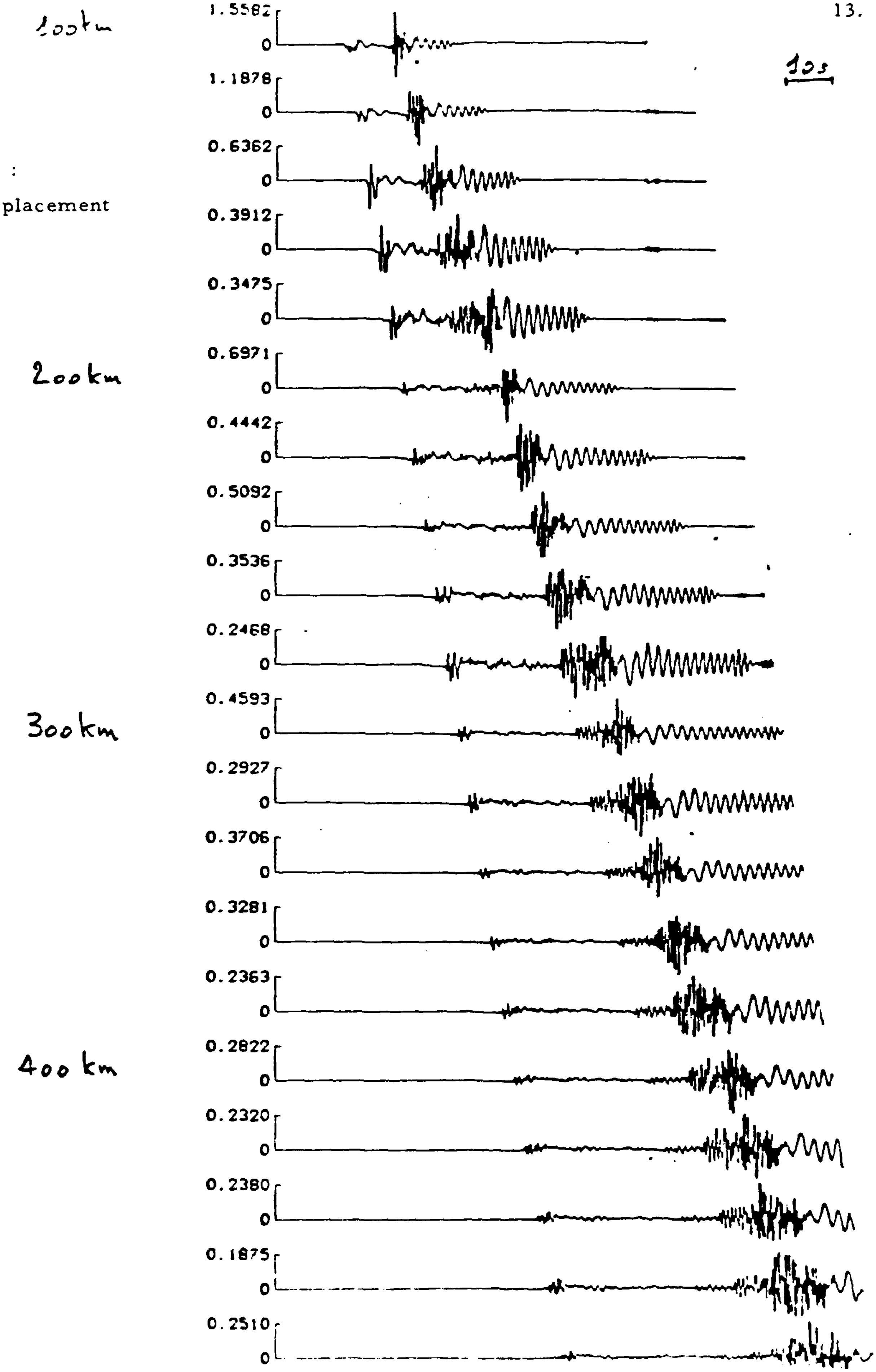
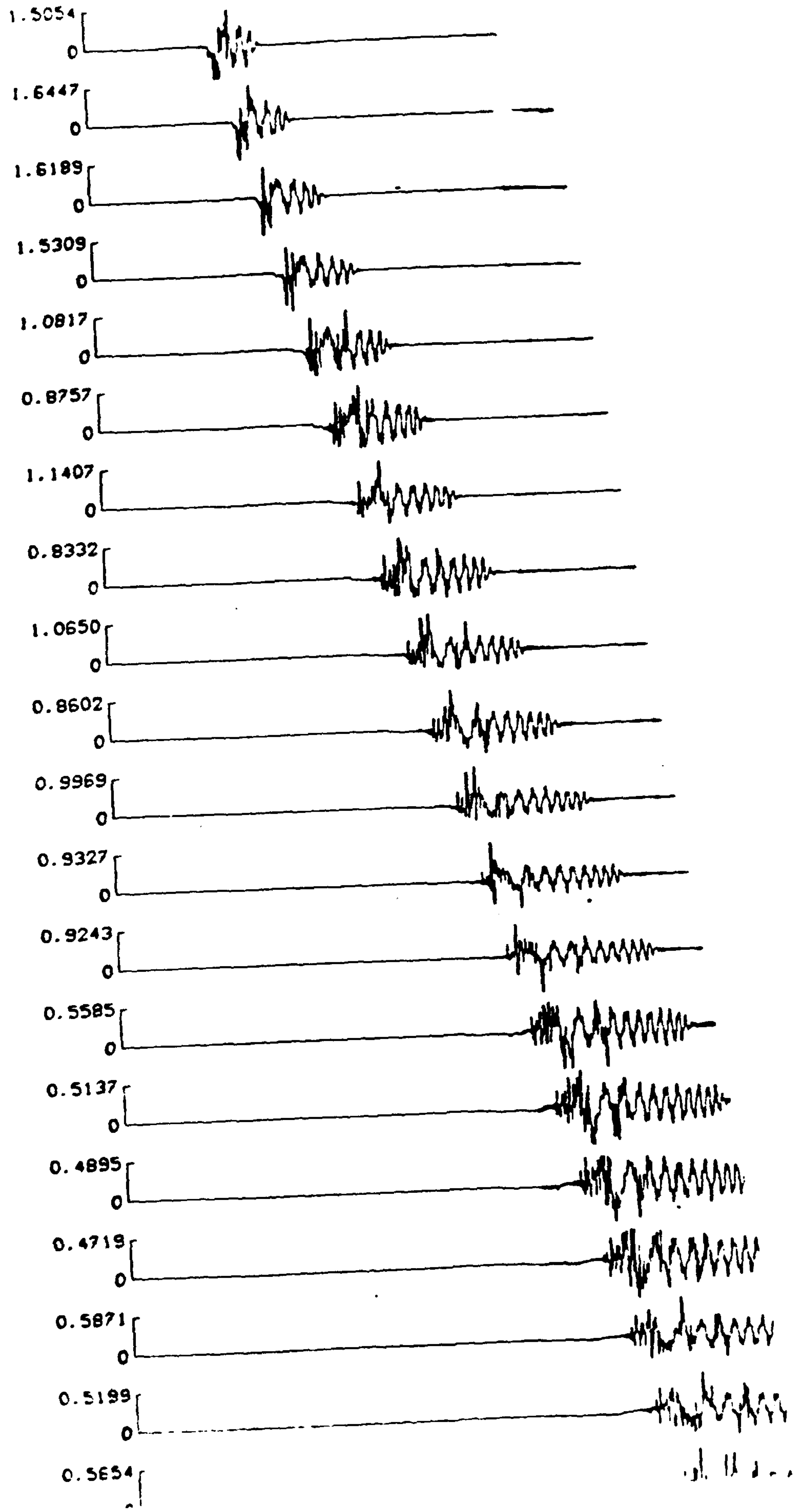


Figure 1b :

Tangential displacement



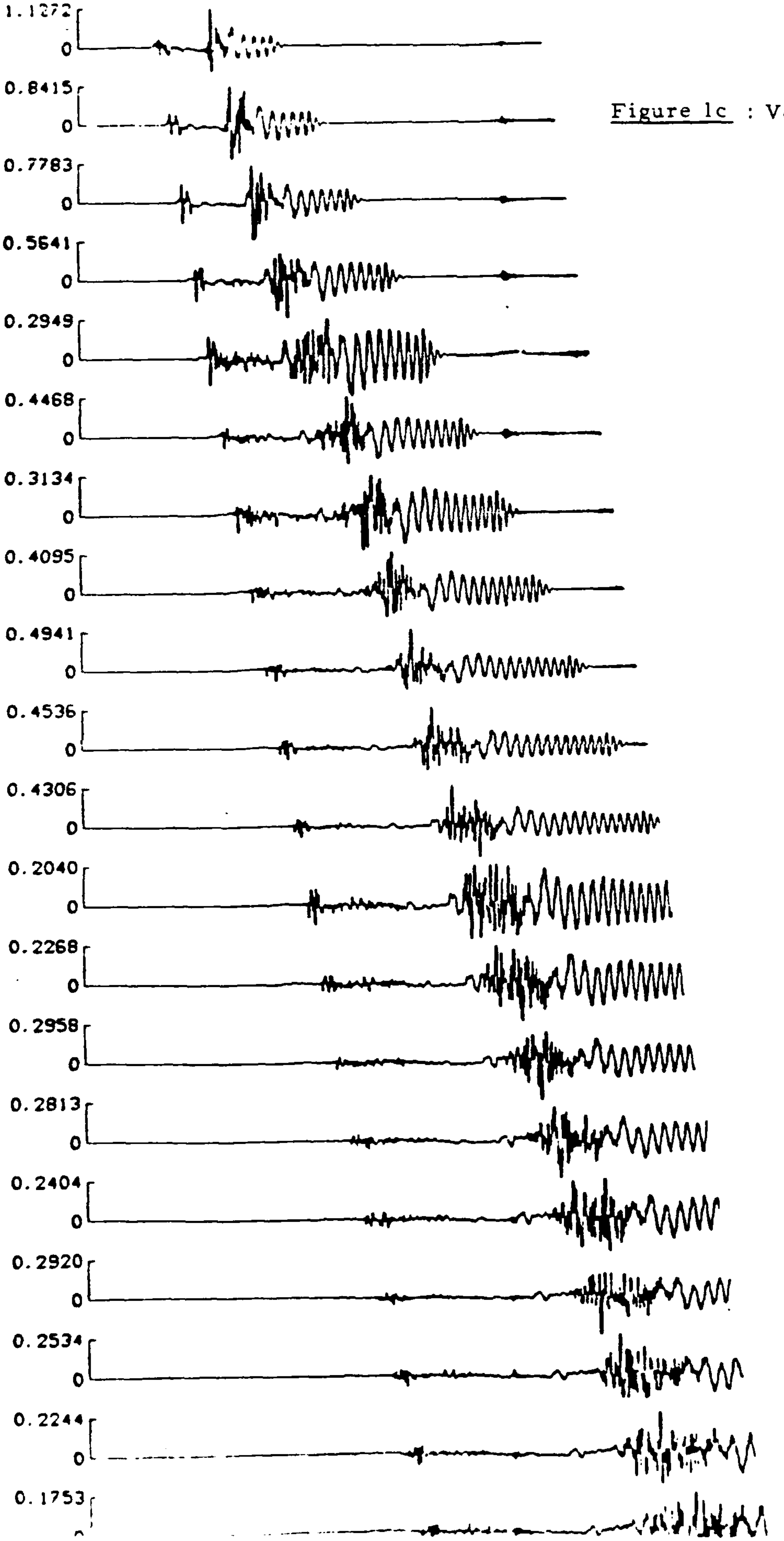


Figure 1c : Vertical displaceme

FIG: 2

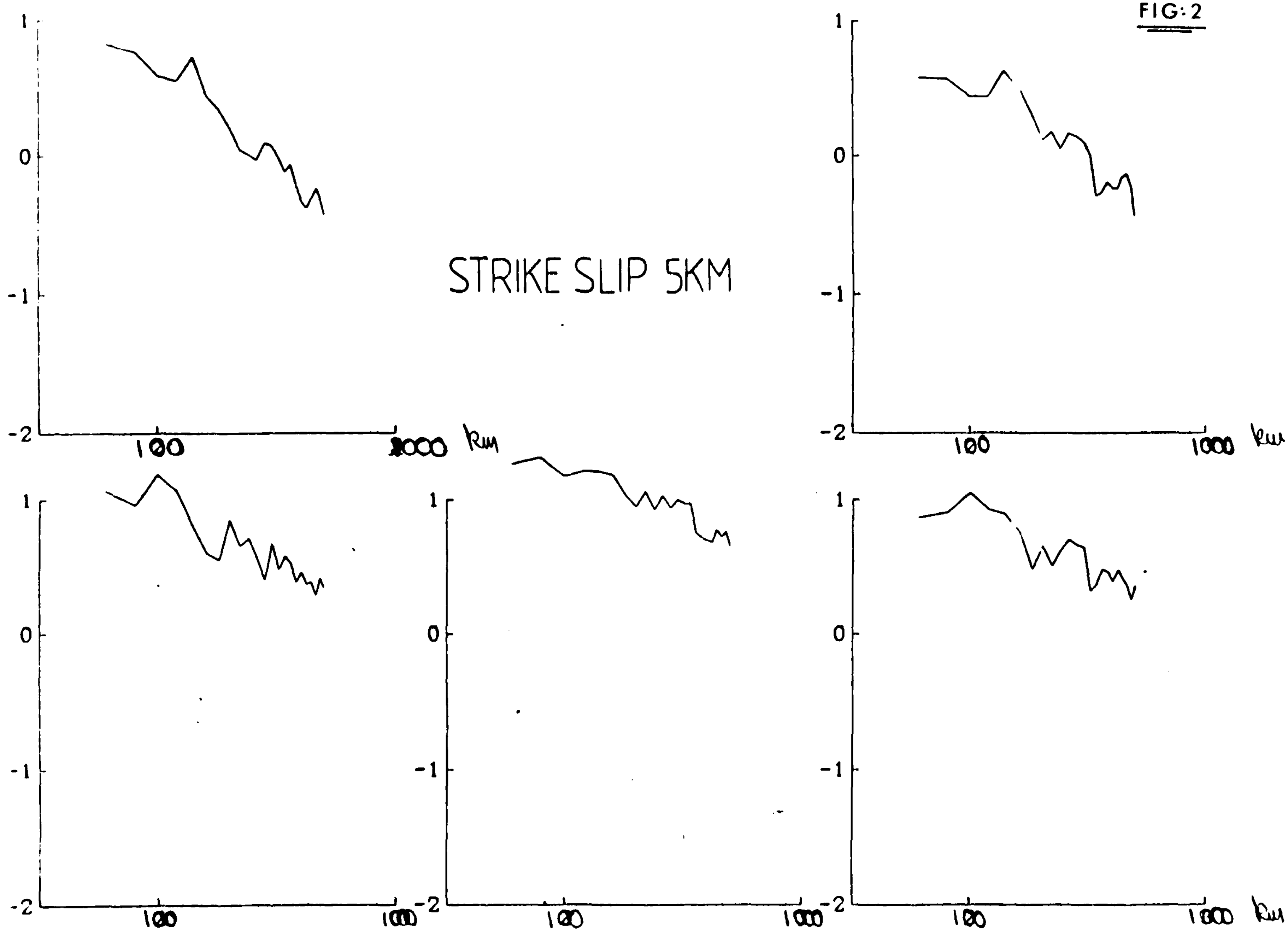
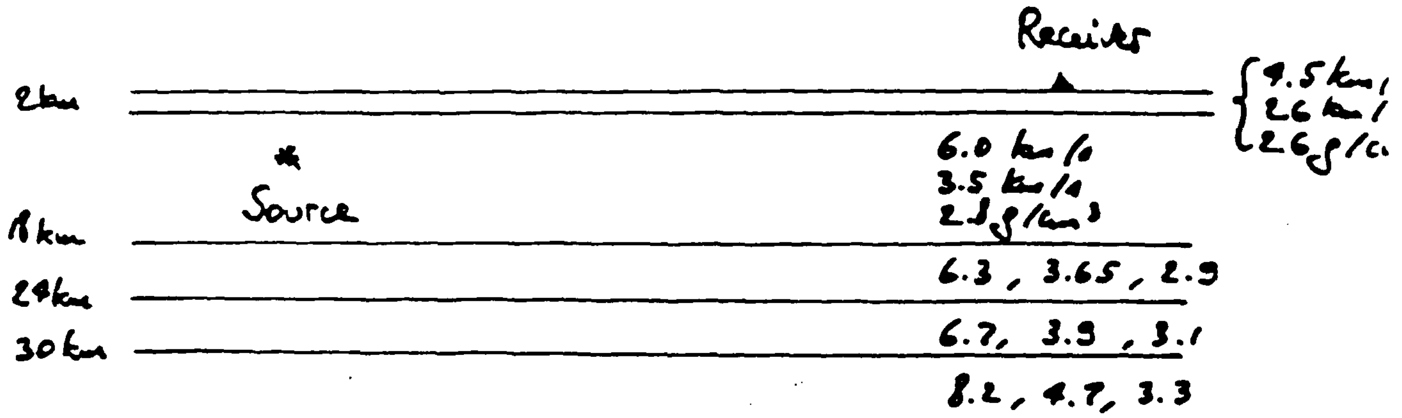


FIG:3



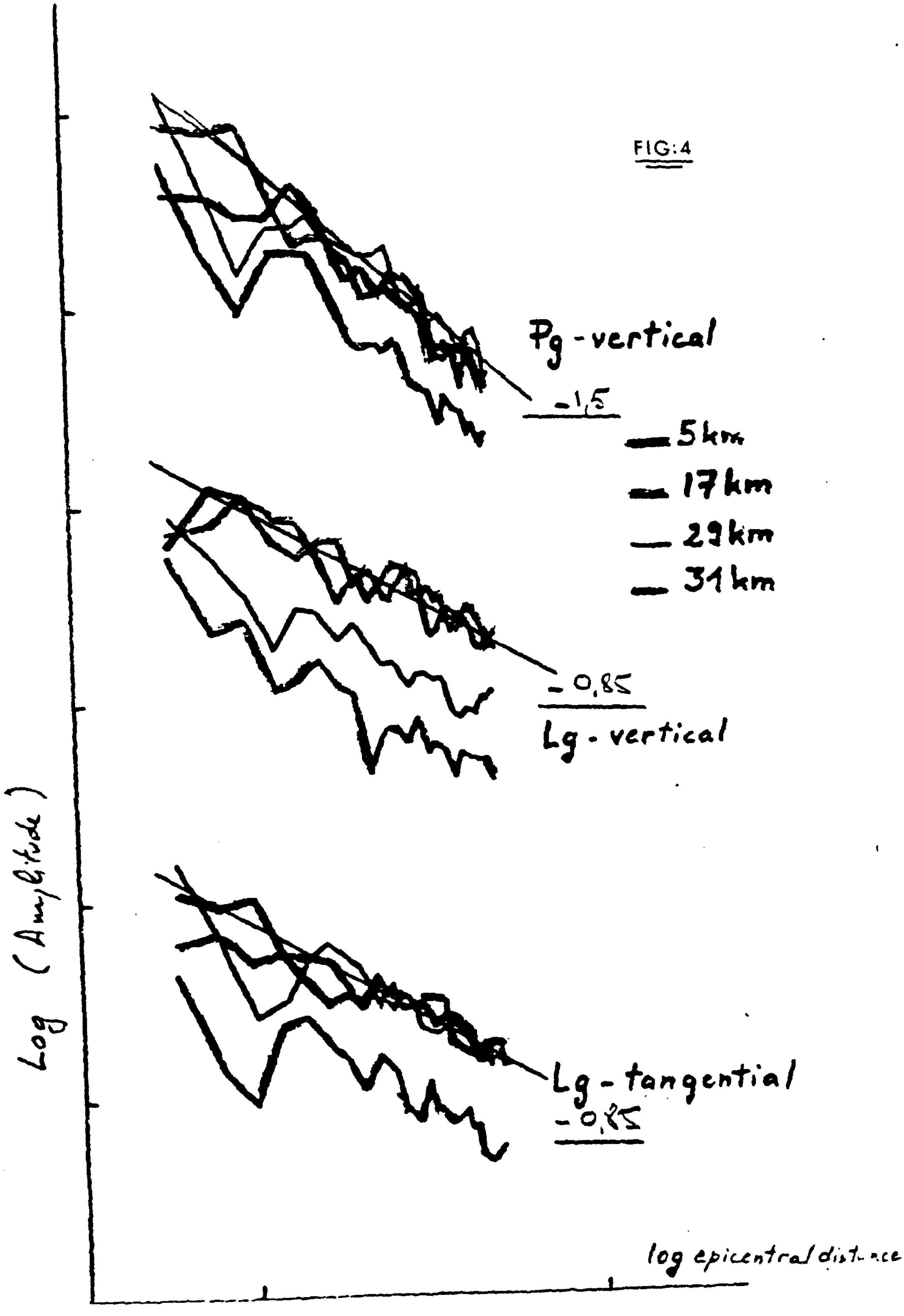


FIG:4
Pg - vertical
-1,5
5 km
17 km
29 km
31 km
-0,85
Lg - vertical
Lg - tangential
-0,85
Log (Amplitude)
log epicentral distance

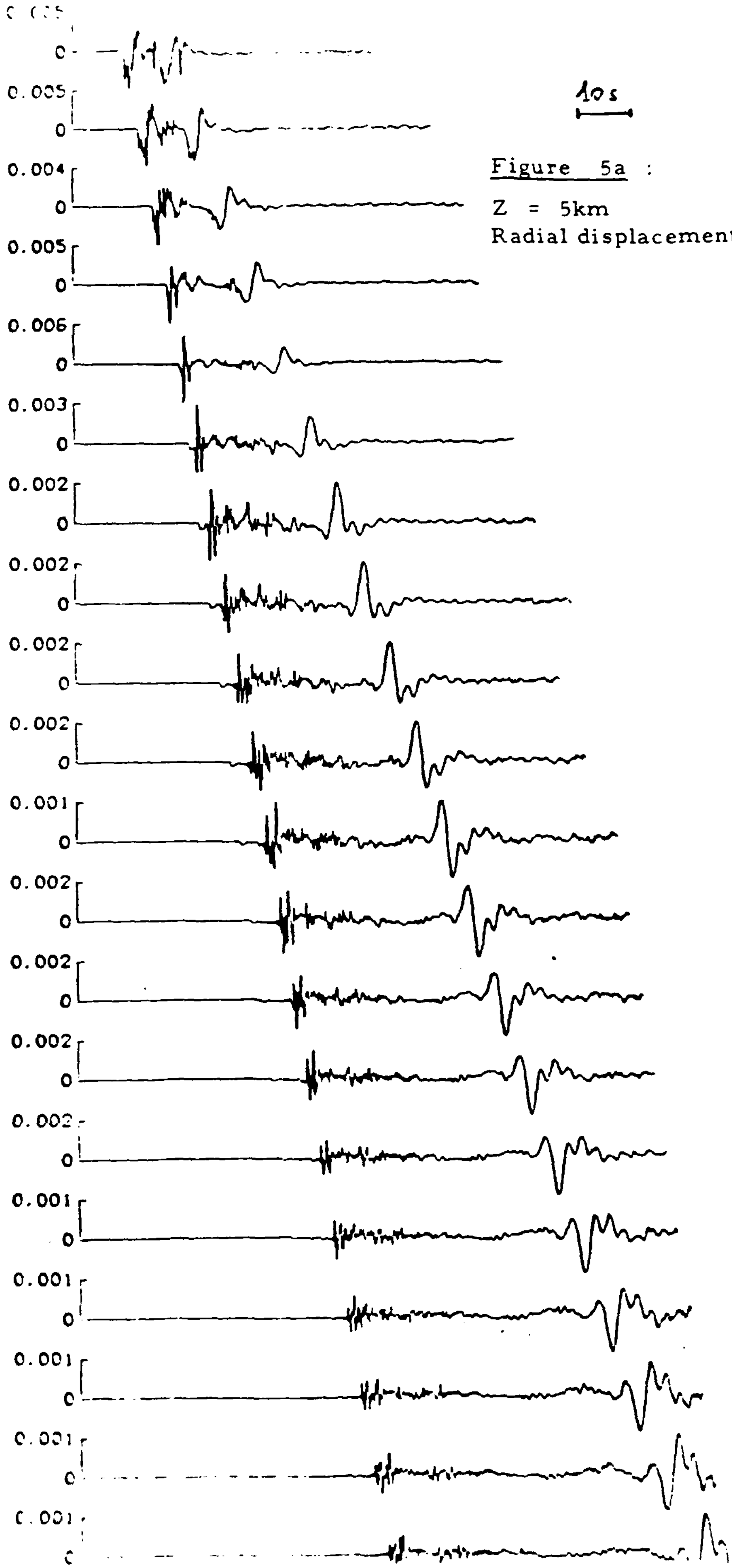


Figure 5a :

Z = 5km
Radial displacement

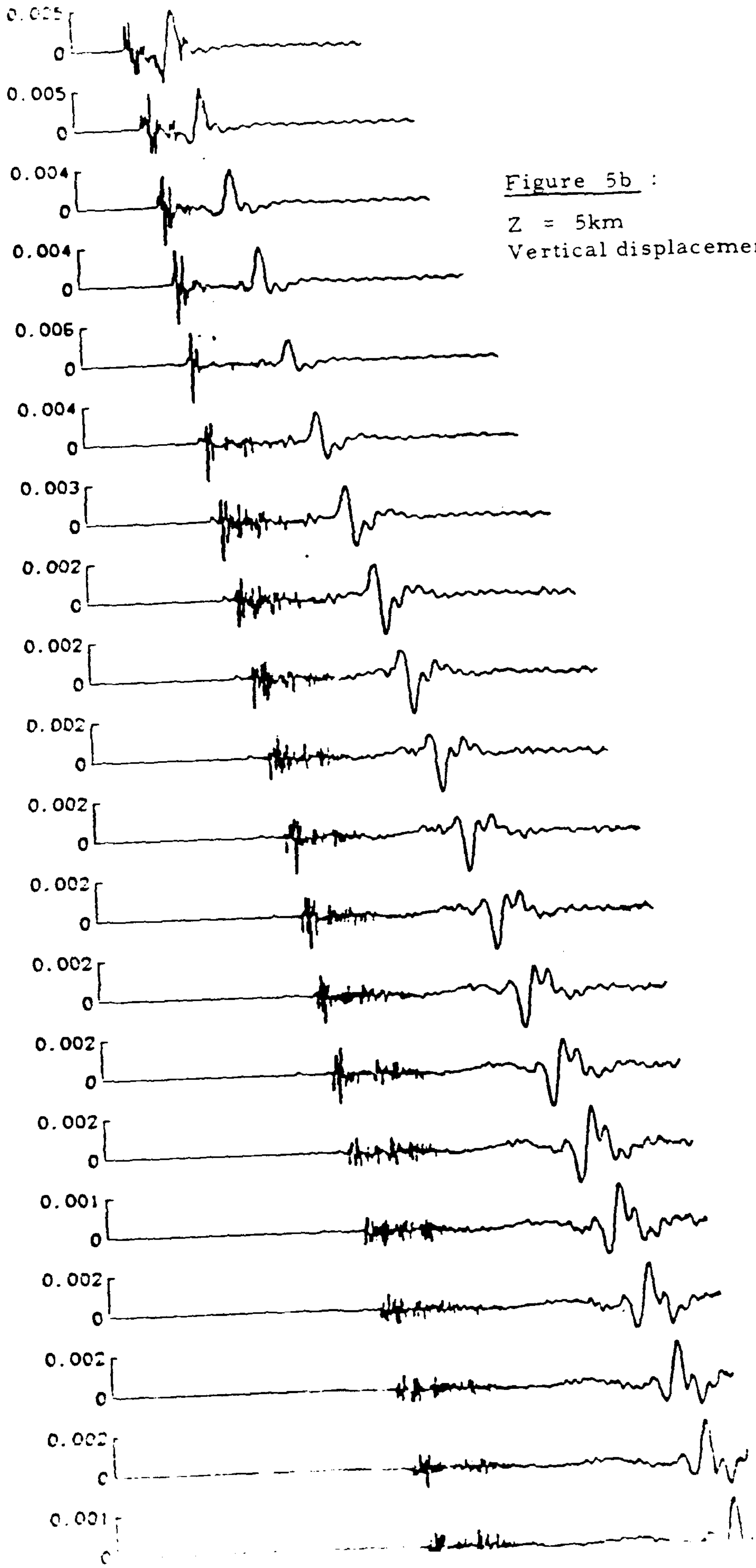


Figure 5b :

Z = 5km
Vertical displacement

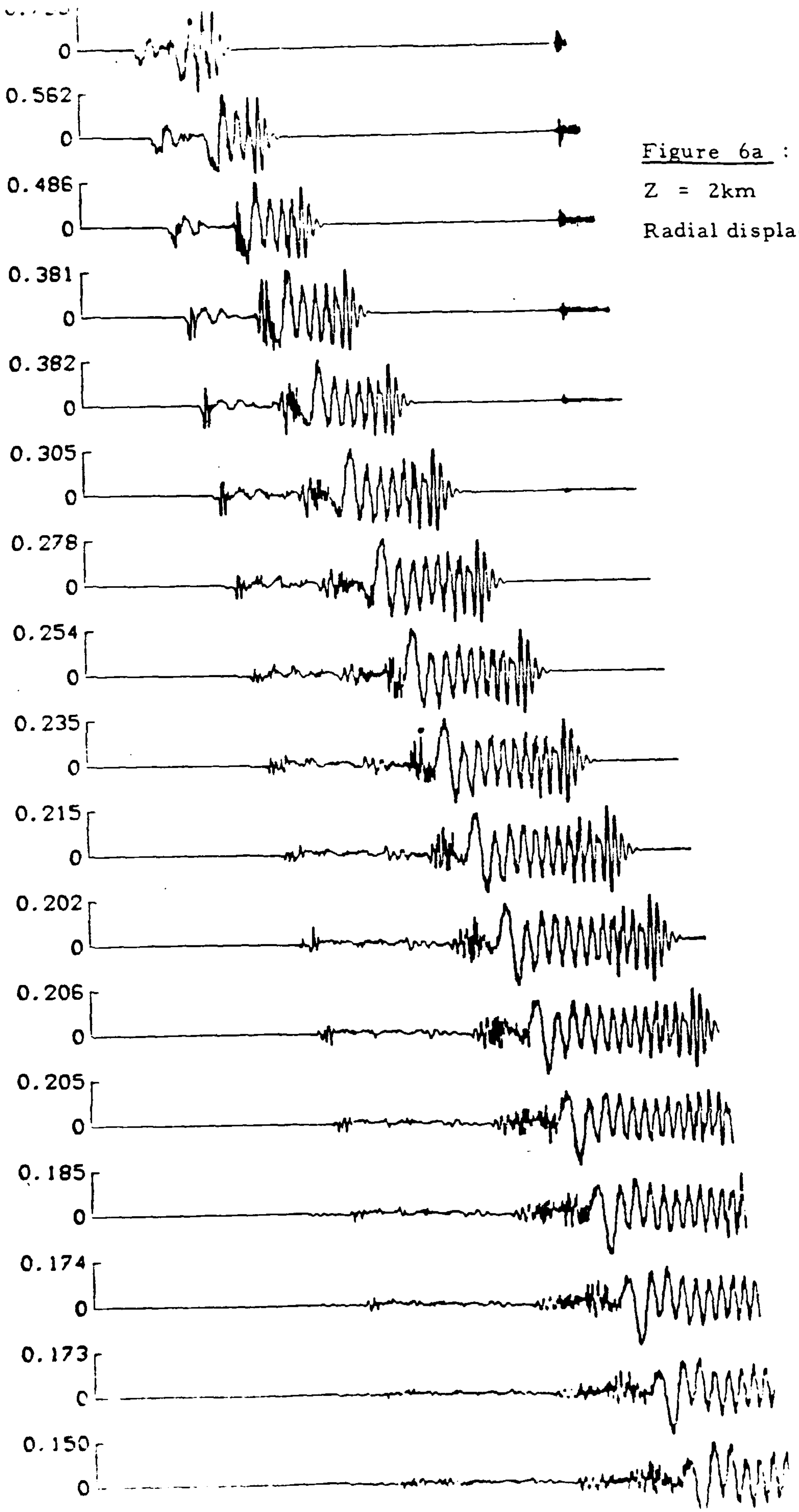


Figure 6a :

Z = 2km

Radial displacement

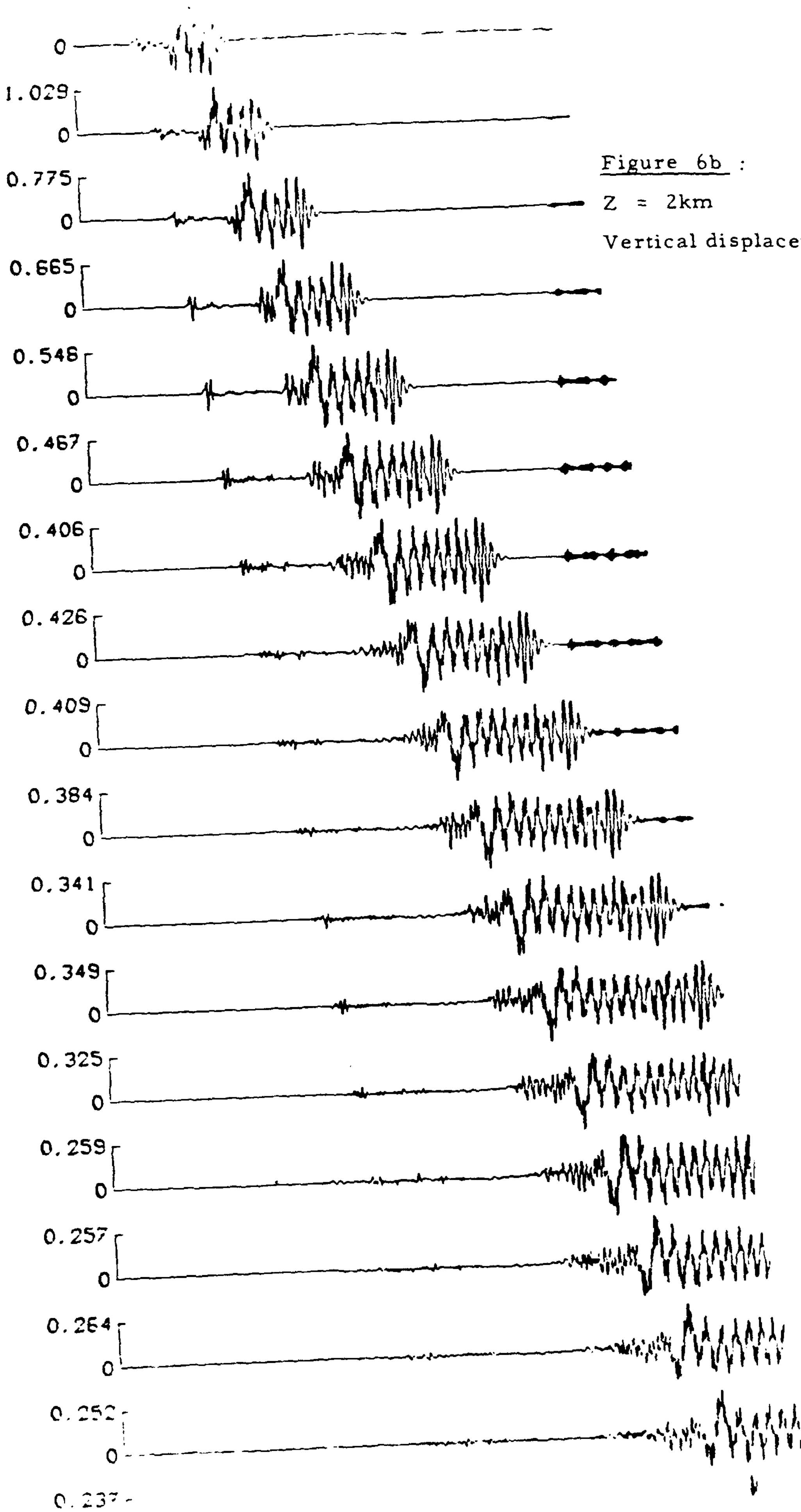


Figure 6b :
Z = 2km
Vertical displacement

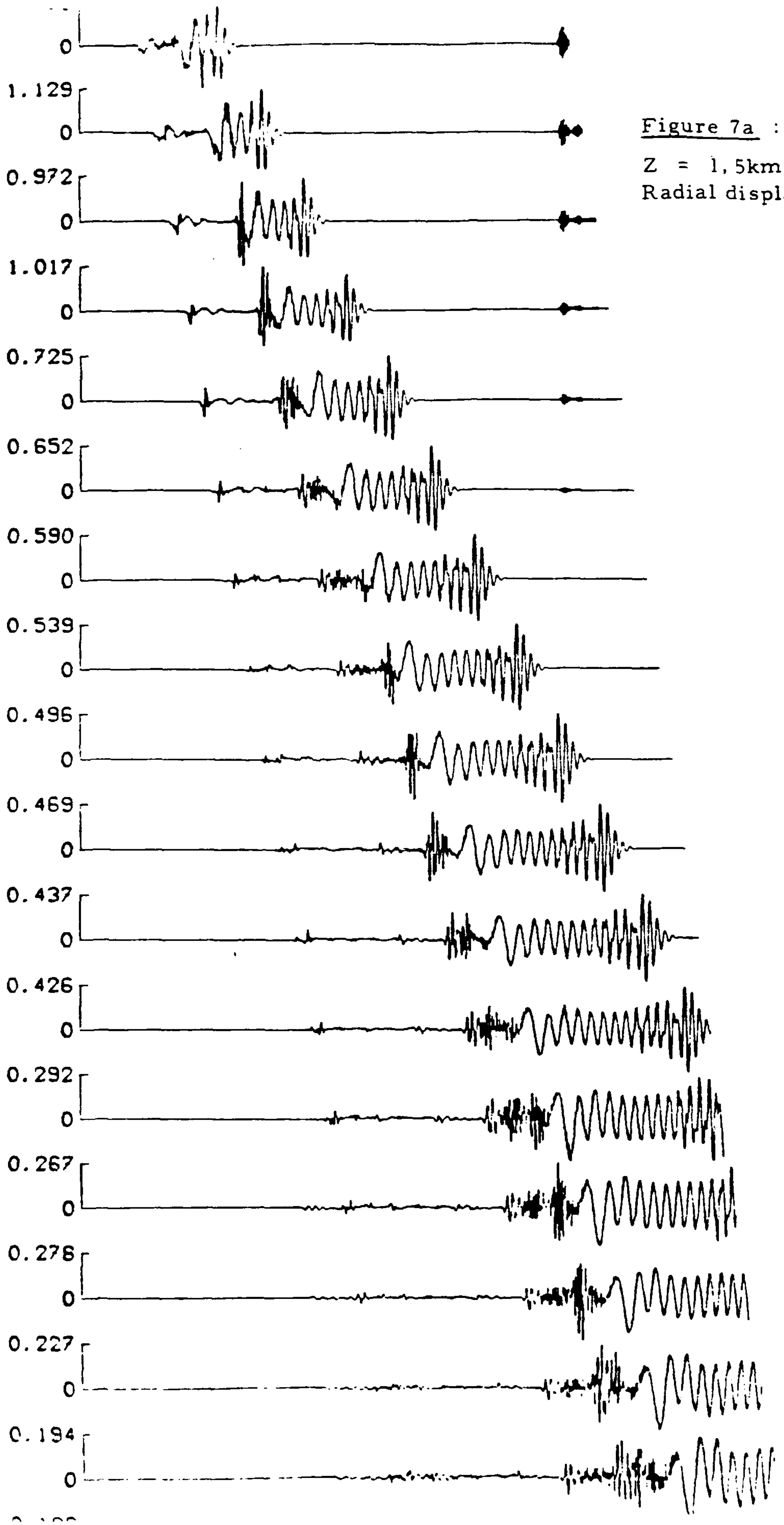
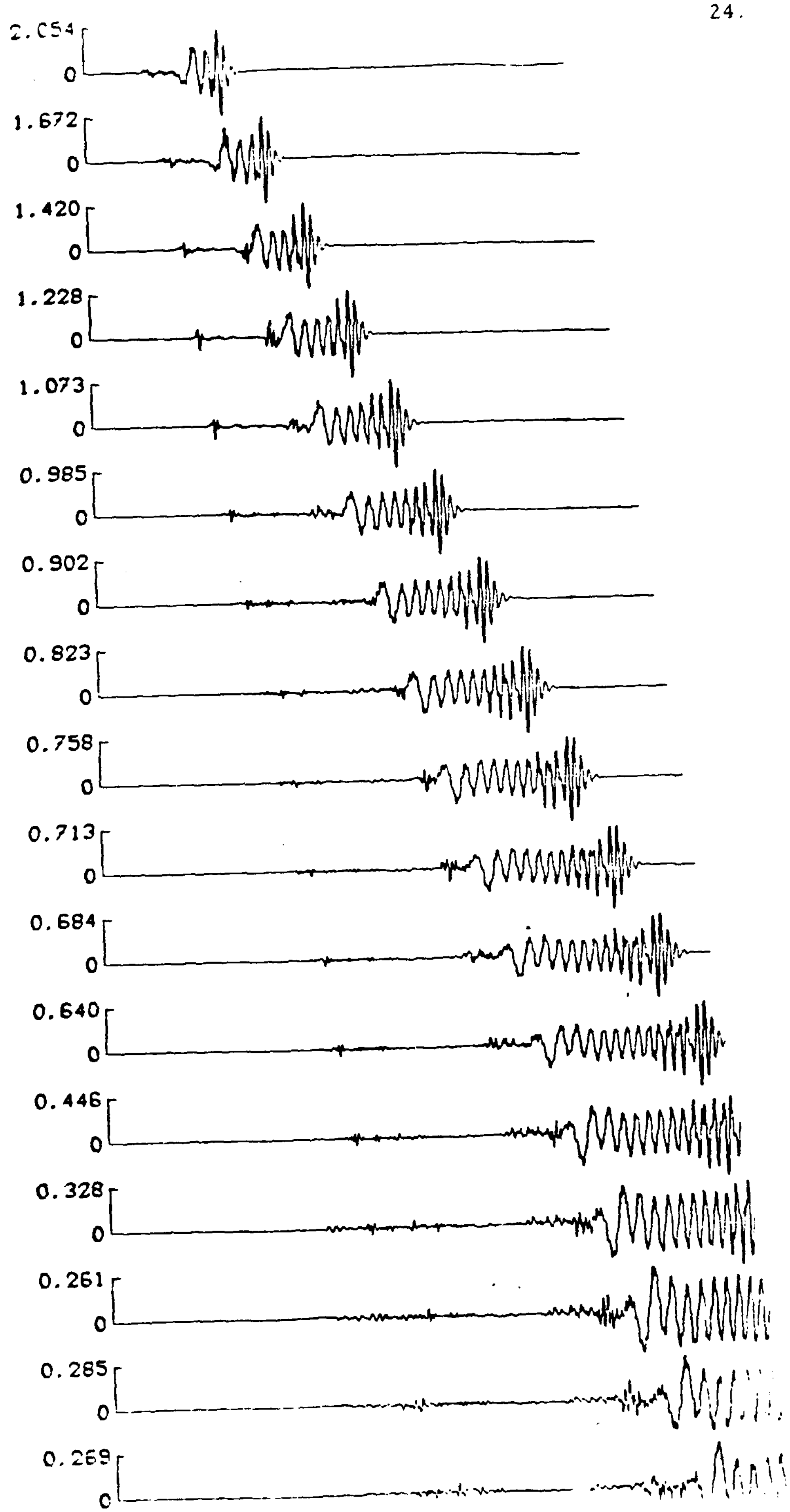


Figure 7a :

Z = 1,5km
Radial displacement

Figure 7b :

Z = 1,5km
Vertical displacement



Propagation of T waves and conversion into seismic waves at a continental slope level

The purpose of the study we have started recently (end of 1982) concerns a better understanding of the T waves we usually record on seismographs on islands, atolls or continental plateau (case of France) and which origin could be an earthquake or an explosion.

In France, the T waves trains recorded for an under sea chemical blast, on continental stations, have a rather long duration : some minutes, and longer is the continental slope, longer is the seismic signal . For similar sources, seismic signals recorded on small islands or atolls in French Polynesia are shorter (some 10 seconds) and with a clear incident wave.

The basic idea which could explain this phenomena is that the T wave signal recorded in one continental seismic station is composed of different simple signals having followed different propagation paths (sofar) in the sea from the source to different conversion points of the continental slope.

At this level, each elementary contact point could be considered as a diffraction source which transmits along different azimuths the energy of the incident T waves.

Before translating this physical approch into some mathematical form , we had first to determine the nature of the converted seismic waves.

Taking the opportunity of some chemical blasts in the Atlantic Ocean at about 1500km from the French continental slope, three componentsdata have been recorded on a SP seismic station set up at 200km from the sources (Fig. 1).

After having applied filtering process as remode filters (rectilinear motion detector) on the data, it was possible to identify the two successive main converted T waves trains as compression and shear waves which propagate at Pg and Sg velocity. In fact, the remode filter is a polarisation filter which computes a cross correlation of the vertical and the supposed radial components versus azimuth to pointed out any P wave reception.

In our case maxima where found inside a $\pm$ 30° angle around the epicentral azimuth. This result seems to confirm a large participation of the continental slope to the wave forms (Fig. 2).

Similar dispersions where found for Sg waves (Fig. 3).
The next step should be to compute for each sample of the continental slope at the sofar depth (1000m), the transfert function of the incident signal to the diffracted signal. Then sum them with reasonable time delay. By convolution with a source function and attenuation function for the crustal propagation, try to rebuilt the whole seismogram at each station which recorded the converted T phase.

If this comprehension of conversion of T phases into seismic waves is satisfactory, we are planning to use then a more complicated source function to simulate an earthquake and see the main differences on the seismograms features.

Another effect to be studied is the conversion between seismic ant T waves at the source level.

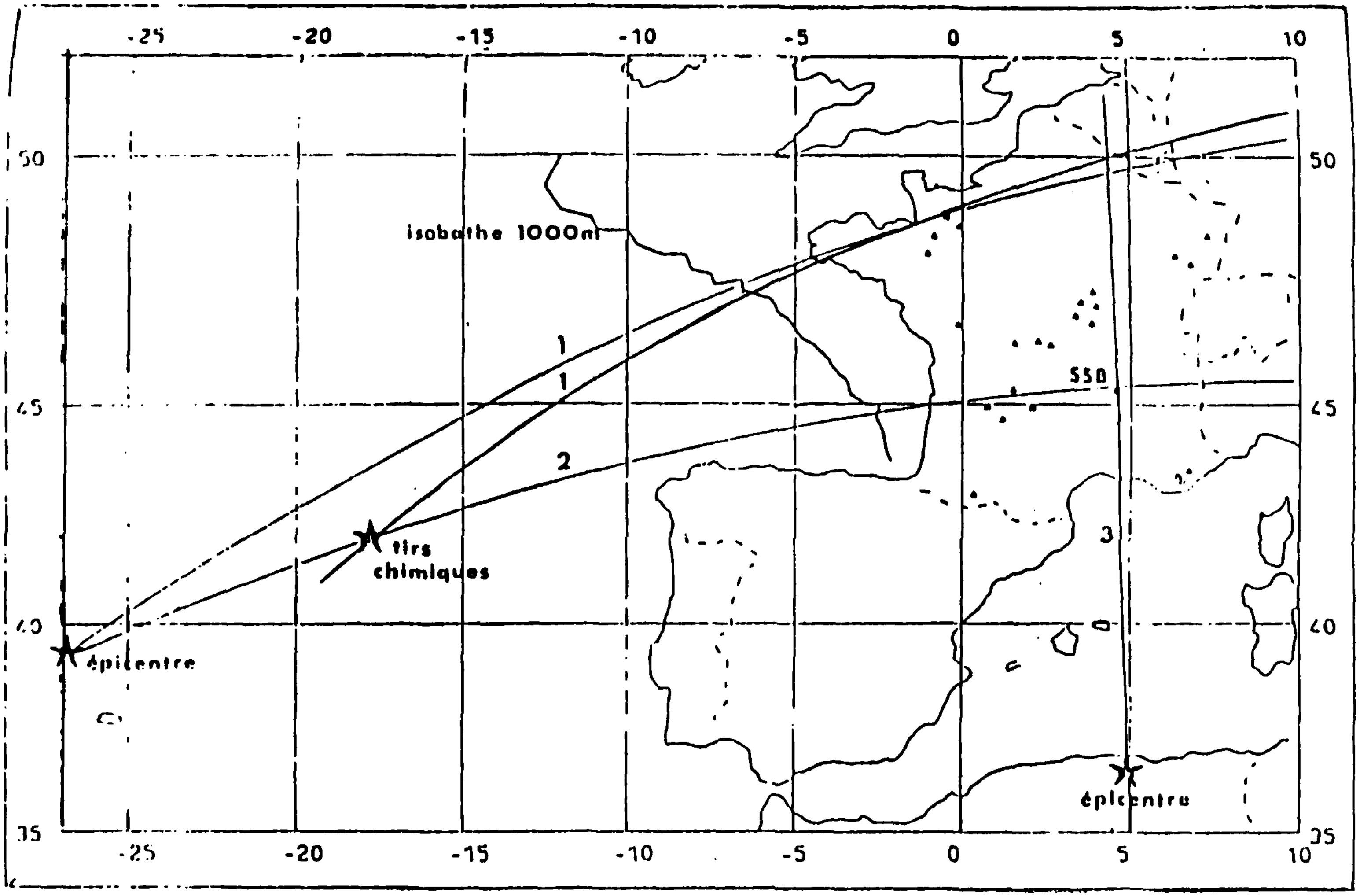


Figure 1 : Ray paths for T and seismic waves.

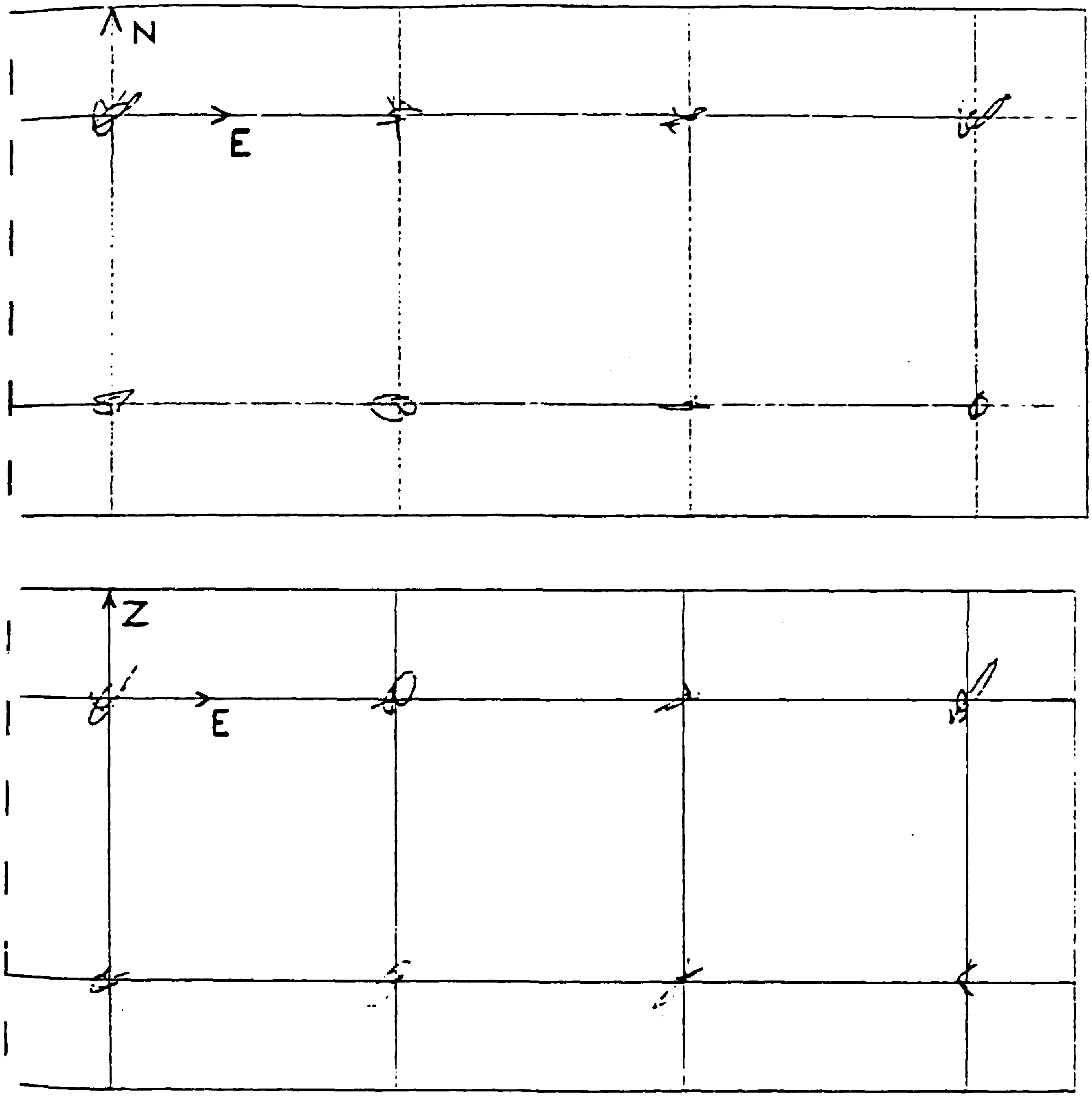


Figure 2 : Displacement in horizontal (above) and vertical (below) plane for P waves.

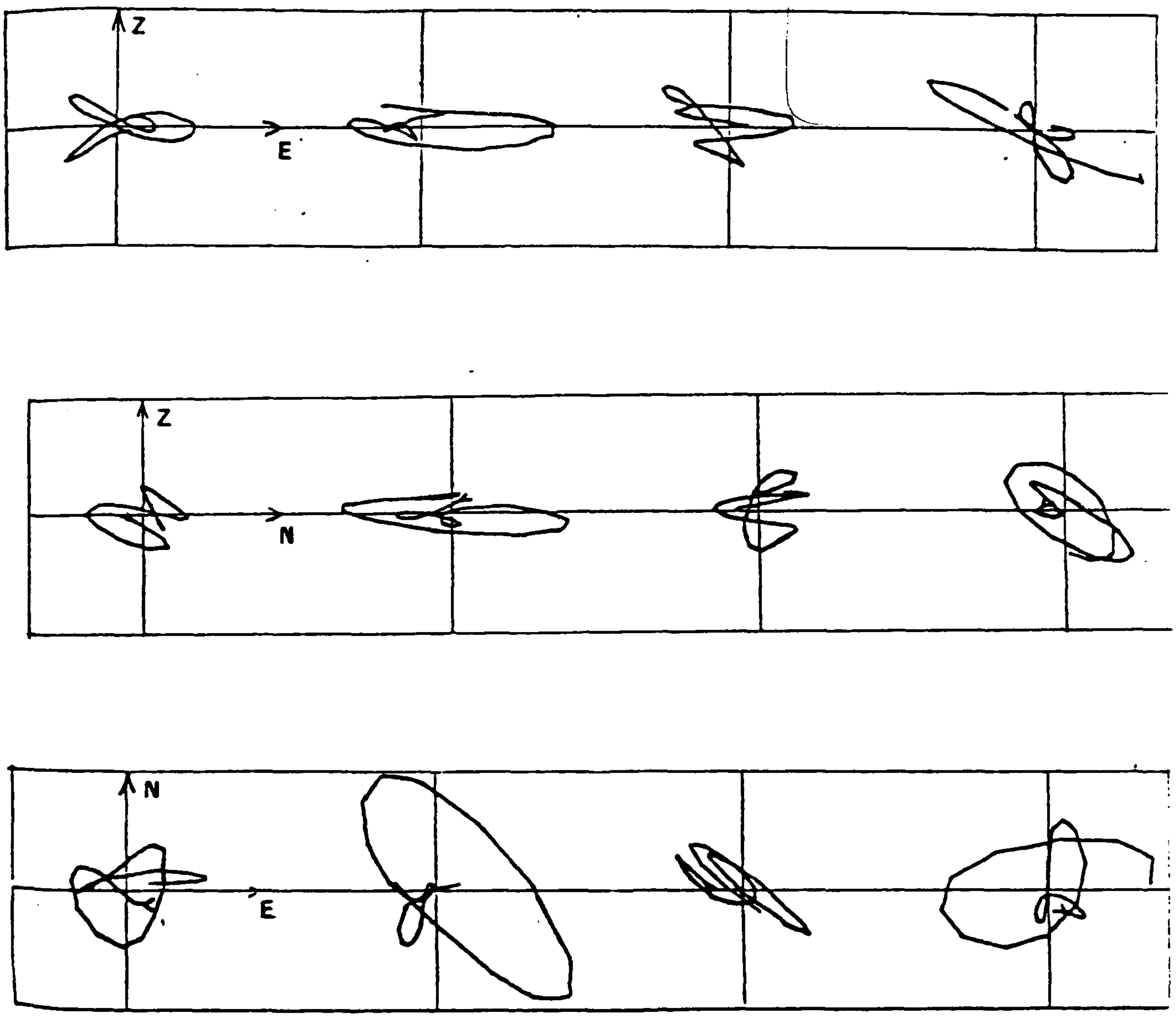


Figure 3 : Displacement in vertical (upper and middle) and horizontal (below) plane.